

Models for the Atmosphere of Mars Based on the Mariner IV Occultation Experiment

by

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RADIOSCIENCE LABORATORY
STANFORD ELECTRONICS LABORATORIES
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ABSTRACT

Several possible atmospheric models are investigated based on data from the radio occultation experiment, and one is shown to be more likely than the others. Profiles in height of the constituent number densities, electron number density, temperature, pressure, and mass density are derived. The analysis indicates that Mars has a tenuous carbon dioxide lower atmosphere with a temperature of about 180 °K near the surface, and an atomic oxygen upper atmosphere with a temperature of only about 80 °K. Frozen carbon dioxide particles may be an almost permanent feature of the atmosphere at intermediate altitudes. The main daytime ionospheric layer has its peak density at 120 km, and is most likely a Bradbury (F2) layer with the principal ion (O^+) being lost through $O^+ + CO_2 \rightarrow O_2^+ + CO$. The atmospheric mass density decreases nearly ten orders of magnitude from the surface to the base of the exosphere at 140 km, thus remaining several orders of magnitude below the density of the earth's atmosphere at corresponding altitudes despite the lower gravity.

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I. INTRODUCTION

The Mariner IV occultation experiment provided new and precise data relative to important properties of the atmosphere of Mars. The experimental results have been presented in several preliminary publications, together with a physical description of several basic properties of the lower atmosphere and the ionosphere [Refs. 1-3]. In addition, initial model studies of the atmosphere are discussed by Fjeldbo et al [Ref. 4] and by Johnson [Ref. 5]. In the present paper we consider in greater detail the possible dominant physical and chemical phenomena in the upper atmosphere and ionosphere in an attempt to derive a more definitive model for the Martian atmosphere. While several possible models are considered, one appears to be more probable than the others based on present information about the critically important rate coefficients for ion loss processes.

The initial results on atmospheric and ionospheric properties as deduced from the occultation experiment were surprisingly different than had been anticipated based on previous work. The model studies, however, indicate that the occultation results for the lower atmosphere and the ionosphere exhibit a degree of self-consistency which tends to substantiate the basic correctness of the interpretations. While much more work should be conducted on data analysis and atmospheric model studies, it seems clear that the following salient characteristics of the atmosphere of Mars have been well established:

1. Mars has a tenuous atmosphere. For example, the molecular number density near the surface is only about 0.8 percent of that of the earth, and corresponds to an altitude of about 34 km on earth.
2. Carbon-dioxide must be the principal atmospheric constituent in order to explain both the occultation and spectroscopic measurements. This suggests that the atmosphere was formed primarily by outgassing of the planet.

In addition, the model studies discussed here suggest that:

3. The carbon dioxide is dissociated between 70 and 80 km by solar ultraviolet radiation, and the CO_2 , CO, and O are distributed in the upper atmosphere by diffusion. Thus atomic oxygen is the principal constituent above about 90 km.

4. The principal ion in the main daytime ionospheric layer (a Bradbury layer, like the F2 region on earth where the profile above the peak is controlled by ambipolar diffusion) is O^+ , and the critical reaction for its loss is probably $O^+ + CO_2 \rightarrow O_2^+ + CO$.
5. The atmosphere of Mars is very cold at all altitudes, the temperature being about 180 °K near the surface (at the time and place of the immersion occultation measurement) and about 80 °K at ionospheric heights. Between these height regions it is even colder, the temperature being at or below the freezing point of CO_2 with frozen CO_2 particles probably being an almost permanent feature of the atmosphere.
6. Because of the low temperature, the atmosphere of Mars is confined near the planet (the exosphere begins near 140 km, where the atomic mean-free-path equals the scale height), and the height profile of atmospheric mass density is several orders of magnitude below that of the earth at all heights even though gravity is 62 percent lower on Mars. Thermal escape of the atmosphere is now insignificant, although loss due to action of the solar wind may play an important role in the continuing evolution of the atmosphere.

The work reported here and the earlier results of Fjeldbo et al [Ref. 4] agree with certain important features of the model developed by Johnson [Ref. 5] which follow from his independent identification of the ionosphere as a Bradbury layer having O^+ as the principal ion. Our work differs from Johnson's, however, with regard to the critically important reaction and rate coefficient for the loss of oxygen ions in the main ionospheric layer, with the result that there are several important differences in our conclusions about the density and temperature profiles. These and other differences are discussed in the final section.

II. THE LOWER ATMOSPHERE

By way of review of previous publications, Table 1 summarizes the results for the lower atmosphere as obtained from the measurements made during immersion of the spacecraft into occultation [Refs. 1-3]. In the table case (a) is for 100 percent CO₂, while case (b) is for 80 to 100 percent CO₂ by number density and the rest is N₂ and Ar in any ratio. While the relative abundances of CO₂, N₂, and Ar are assumed, all of the other results shown in the table follow directly from the atmospheric refractivity and scale height measured near the surface of Mars. The predominance of carbon dioxide in these models was chosen because the

TABLE 1. SUMMARY OF PRELIMINARY RESULTS
FOR THE LOWER ATMOSPHERE OF MARS

1. Location	Immersion was over Electris at 50 deg S latitude, 177 deg E longitude, local time 1 p.m., winter, solar zenith angle 67 deg
2. Surface Refractivity	3.6 ± 0.2 N-units
3. Surface Scale Height	9 ± 1 km
4. Surface Number Density	
(a) all CO ₂	$1.9 \pm 0.1 \times 10^{17}$ mol/cm ³
(b) up to 20 percent N ₂ and/or Ar	$2.1 \pm 0.2 \times 10^{17}$ mol/cm ³
5. Surface Mass Density	
(a) all CO ₂	$1.43 \pm 0.10 \times 10^{-5}$ g/cm ³
(b) up to 20 percent N ₂ and/or Ar	$1.50 \pm 0.15 \times 10^{-5}$ g/cm ³
6. Temperature Near Surface	
(a) all CO ₂	180 ± 20 °K
(b) up to 20 percent N ₂ and/or Ar	175 ± 25 °K
7. Surface Pressure	
(a) all CO ₂	4.9 ± 0.8 mb
(b) up to 20 percent N ₂ and/or Ar	5.1 ± 1.1 mb

total surface pressure obtained from the occultation experiment is then in best agreement with spectroscopic observations of Mars [Refs. 6-9].

The main uncertainty in these results is the height of the occulting surface feature above the mean surface. Another source of error which has been suggested is dust suspended in the atmosphere. These two effects would act in opposite directions, but only the altitude effect, amounting to a 1 percent change in number density per 90 meters of height, is thought to be of possible significance.

Ionization near the surface of Mars could also cause errors. However, even during abnormal blackout in the earth's polar atmosphere, the electron number density apparently does not exceed 10^3 cm^{-3} at the density level corresponding to the Martian surface conditions [Ref. 10]. Using 10^3 cm^{-3} as an upper limit to the surface electron density at Electris during the measurement, one finds that this effect would increase the number density given in the table by less than 0.3 percent.

The number density scale height deduced from the immersion data appears to be constant with altitude at least up to about 30 km [Refs. 1-3]. This result seems reasonable also on theoretical grounds, since radiative exchange considerations show that the temperature lapse rate in a carbon dioxide atmosphere should be relatively low in winter at high latitudes [Ref. 11], which corresponds to conditions during immersion.

The emersion measurements over Mare Acidalium in the northern hemisphere of Mars indicate a surface pressure which is several millibars higher than the immersion results. The difference in surface pressure may be caused by an altitude difference, due either to local surface features or to departures of the mean surface from a gravitational equipotential surface. The temperature and apparent lapse rate were also higher over Mare Acidalium. This is to be expected since it was summer in the northern hemisphere during the measurement.

III. THE UPPER ATMOSPHERE

From the immersion measurements over Electris, the preliminary electron number density profile of Fig. 1 has been derived. The local time was 1300 and the solar zenith angle was about 67 deg during the measurement. The data and the computational techniques used to obtain this profile have been published previously [Refs. 1-3, 12].

The main ionospheric layer is at an altitude of 120 ± 5 km and has a peak electron density of $9.5 \pm 1 \times 10^4 \text{ cm}^{-3}$. The plasma scale height at the top side of this layer is 24 ± 3 km. Both the small scale height and the low altitude of the main layer show that the atmosphere is substantially colder than previously anticipated.

There is also evidence of a minor ionospheric layer at approximately 100 km altitude. The electron number density in this layer is about $2.5 \times 10^4 \text{ cm}^{-3}$.

While the accuracy of the electron density at the peak of the main layer is of the order of 10 percent, that portion of the profile having lower density is less accurate. In addition, nonspherical trends in the electron density distribution cause increasing errors towards the lower portion of the bottom side of the ionosphere.

Measurements were also made as the radio energy probed through the midnight ionosphere above Mare Acidaliu. However, the S-band frequency is not very sensitive to the low electron number densities encountered in this region [Refs. 4, 13], and at present we can only say that the peak electron density on the night side was probably less than $5 \times 10^3 \text{ cm}^{-3}$. This upper limit for the midnight density together with the peak electron density measured on the day side, both tend to suggest a somewhat higher loss coefficient than is found in the earth's F2 region.

We will now briefly discuss some of the physical and chemical processes thought to be important in the upper atmosphere of Mars.

Solar ultraviolet radiation shortward of about 1700 \AA dissociates carbon dioxide [Refs. 14, 15]:



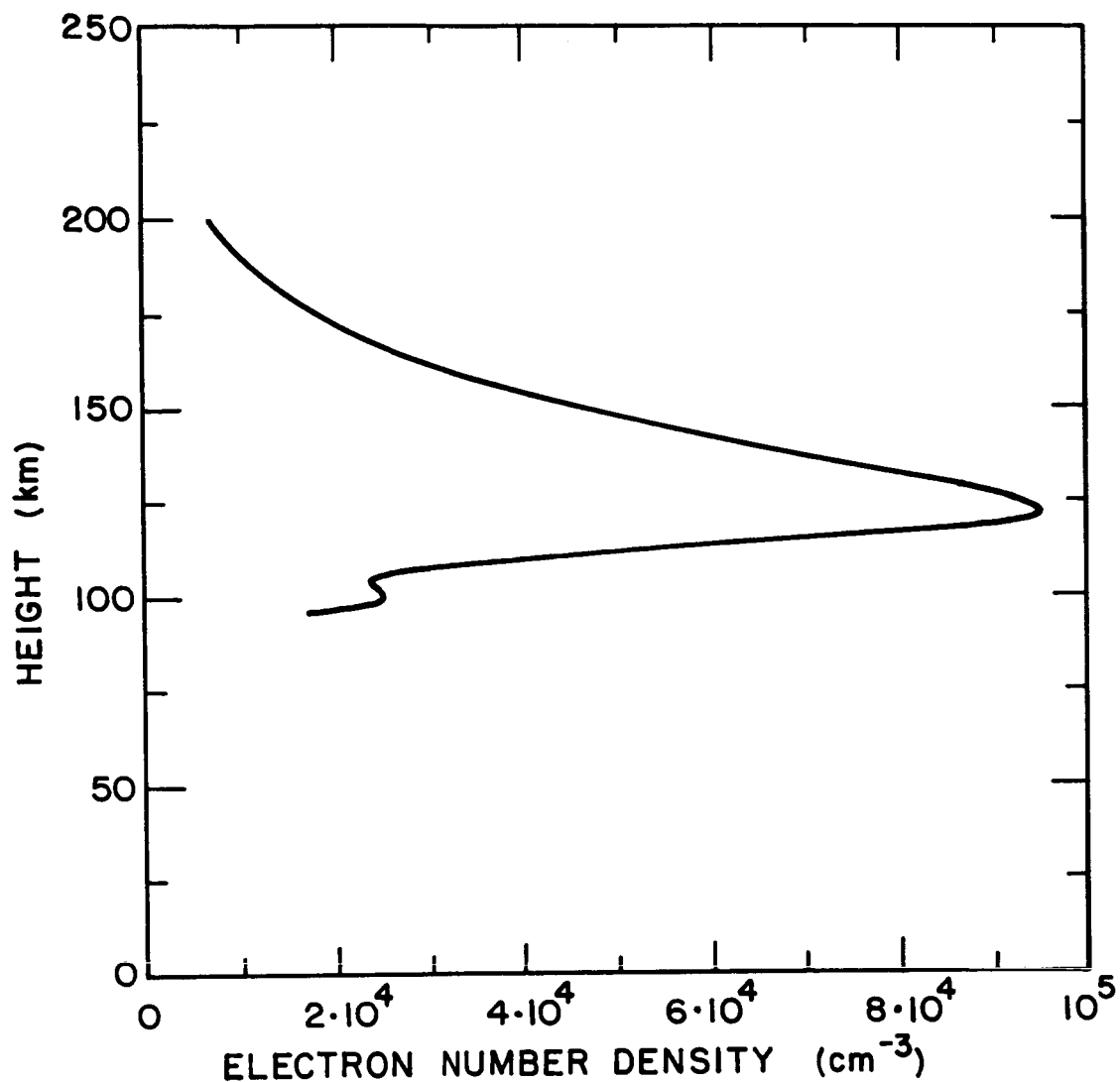


FIG. 1. ELECTRON NUMBER DENSITY VS ALTITUDE ABOVE ELECTRIS AT THE TIME AND LOCATION CORRESPONDING TO IMMERSION INTO OCCULTATION (50 DEG S LATITUDE, 177 DEG E LONGITUDE, 1300 LOCAL TIME, LATE WINTER, SOLAR ZENITH ANGLE 67 DEG).

Association can take place by three-body collisions:



where M is any atom or molecule. Some of the oxygen atoms will also associate:



Molecular oxygen is again dissociated by radiation in the Schumann-Runge continuum:



Similar equations apply for the formation and loss of ozone.

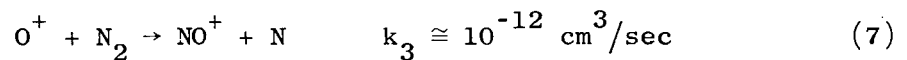
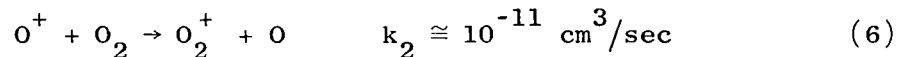
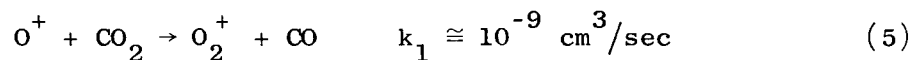
In order to determine the altitude distribution of the different constituents, it is also necessary to consider diffusion. At ionospheric heights, the atmosphere is so rarefied that molecular diffusion becomes the fastest process and diffusive separation is expected to prevail. Hence, above some altitude one may therefore expect that atomic oxygen should become the principal constituent because it is lighter than CO, N₂, O₂, Ar and CO₂.

The Martian atmosphere is exposed to various ionizing agents. While Mars has little or no magnetic field to shield it from corpuscular radiation, solar ultraviolet and X-rays still appear to be the most important ionizing agents except, perhaps, during disturbed solar conditions [Ref. 15].

As we discuss below, several of the preliminary models we have made for the Martian atmosphere have O⁺ as the principal ion in the main ionospheric layer. In these models, the temperature at the top side of the ionosphere would need to be about 80 °K to correspond with the electron scale height of about 24 km.

The crucial problem in determining characteristics of the upper atmosphere, assuming that atomic oxygen is the principal constituent, is the identification of the loss mechanism for O^+ at the peak of the layer. We will consider several possibilities for this "bottleneck" reaction.

It has been shown by Bates that the rate coefficient for radiative recombination of O^+ with electrons increases with decreasing temperature [Ref. 10]. Nevertheless, it is not possible to explain the ion loss rate in the Martian ionosphere on this basis because radiative recombination proceeds too slowly by about 3 orders of magnitude. We must therefore assume that minor constituents take part in the loss process. The most likely possibilities appear to be:



At 120 km altitude, the charge of molecular ions is lost quite rapidly through dissociative recombination with electrons ($O_2^+ + e \rightarrow O + O$ and $NO^+ + e \rightarrow N + O$).

Reaction 5, brought to our attention by C. G. Little and E. E. Ferguson, should be considered because of the preponderance of CO_2 in the lower atmosphere of Mars. By analogy with the earth's ionosphere, one may also expect that the two last reactions could play a role in the loss of O^+ [Ref. 10].

In models where the principal ion is O^+ and any one of the above reactions is the predominant loss process, the electron density profile would take on the form of a Bradbury (F2) layer. Thus with these models for the upper atmosphere of Mars, the electron production peak is below the electron density peak, and the profile on the top side of the layer is controlled by ambipolar diffusion. From the theory for such a layer

[Ref. 16], it is possible to determine the number density of atomic oxygen at the altitude (120 km) of the maximum of the electron density, as indicated below.

Balancing production and loss of O^+ at the peak of the layer yields:

$$\sum_{\lambda} \sigma_{\lambda} I_{\lambda} n(O) = n(O^+) \beta \quad (8)$$

where

σ_{λ} = the ionization cross section of atomic oxygen for different wavelengths

I_{λ} = the intensity of the ionizing flux

$$\sum_{\lambda} \sigma_{\lambda} I_{\lambda} \cong 10^{-7} \text{ sec}^{-1} \quad [\text{Ref. 10}]$$

$n(O)$ = the atomic oxygen number density

$n(O^+)$ = the number density of atomic oxygen ions

β = the loss coefficient for atomic oxygen ions.

If we also equate diffusion time (H^2/D) and loss time ($1/\beta$) at the peak of the layer we obtain:

$$\beta = D/H^2 \quad (9)$$

where the diffusion coefficient (D) is related to the temperature (T) and density [$n(O)$]

$$D = b\sqrt{T}/n(O) \quad (10)$$

where b is a constant, approximately equal to $10^{17} \text{ cm}^{-1} \text{ sec}^{-1} \text{ } ^\circ\text{K}^{-1/2}$ [Ref. 16].

Eliminating β and solving Eq. (8) for $n(O)$ gives:

$$n(O) = \sqrt{\frac{n(O^+) b \sqrt{T}}{H^2 \sum_{\lambda} \sigma_{\lambda} I_{\lambda}}} \quad (11)$$

Equating $n(O^+)$ to the measured peak electron density, and using the temperature (80 °K) and neutral scale height (12 km) derived from the profile measurements, we calculate an atomic oxygen number density of approximately 10^9 cm^{-3} at 120 km altitude. This oxygen density is independent of which of the reactions (5), (6) or (7) is the predominant loss mechanism for O^+ .

IV. THE ATOMIC OXYGEN - CARBON DIOXIDE MODEL FOR THE UPPER ATMOSPHERE

At this time, the most promising atmospheric model appears to be one in which reaction (5) is the dominant loss mechanism in the main ionospheric layer [Ref. 4].

The loss coefficient for this reaction is given by:

$$\beta = k_1 n(\text{CO}_2) \quad (12)$$

Utilizing Eqs. (8) and (12), we obtain an expression for the CO_2 number density required to balance production of O^+ with the loss given by reaction (5):

$$n(\text{CO}_2) = \frac{n(\text{O}) \sum_{\lambda} \sigma_{\lambda} I_{\lambda}}{k_1 n(\text{O}^+)} \quad (13)$$

From laboratory measurements, the rate coefficient k_1 has been found to be approximately $10^{-9} \text{ cm}^3/\text{sec}$ at 300 °K [Ref. 17]. Thus, assuming this value for k_1 , there would be only about 10^6 CO_2 molecules per cubic centimeter at 120 km for the atomic oxygen - carbon dioxide model of the upper atmosphere.

The number density profiles, the temperature profile, and the pressure profile for this model are shown in Figs. 2, 3, and 4, respectively. Below 30 km altitude, the CO_2 number density, $n(\text{CO}_2)$, was calculated from the refractivity measured in the lower atmosphere. At ionospheric heights, $n(\text{O})$ and $n(\text{CO}_2)$ were determined from the observed electron density profile together with ion production and loss considerations as outlined above.

The CO_2 number density profile at ionospheric heights, when calculated in this manner, is inversely proportional to the rate coefficient k_1 . It should be kept in mind that k_1 was measured at 300 °K rather than at 80 °K, and that the temperature dependence is not well known.

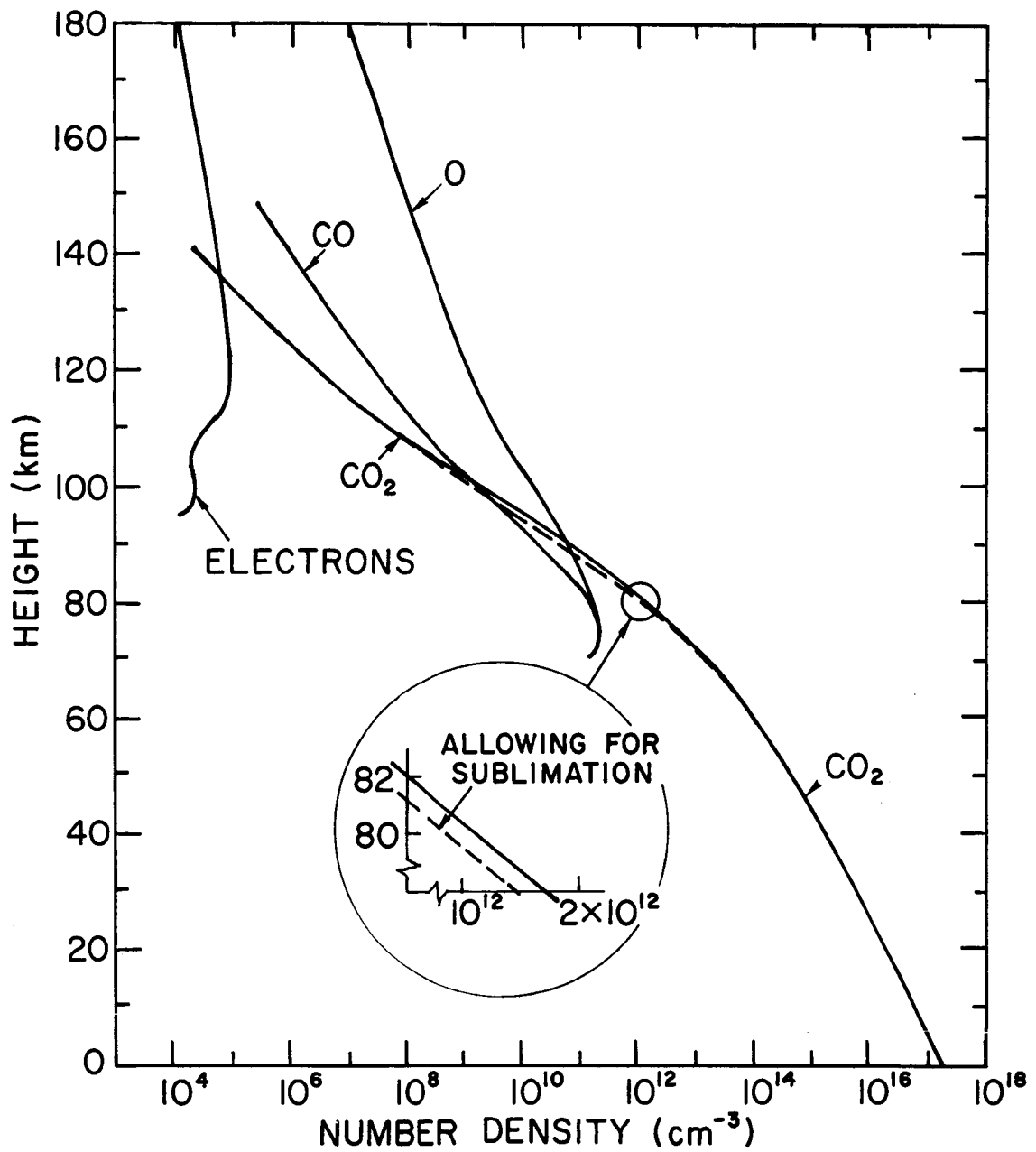


FIG. 2. NUMBER DENSITY VS ALTITUDE ABOVE ELECTRIS AT THE TIME OF IMMERSION INTO OCCULTATION. The limiting ion loss process in the main ionospheric layer is assumed to be $O^+ + CO_2 \rightarrow CO + O_2^+$ with a rate coefficient of 10^{-9} cm³/sec.

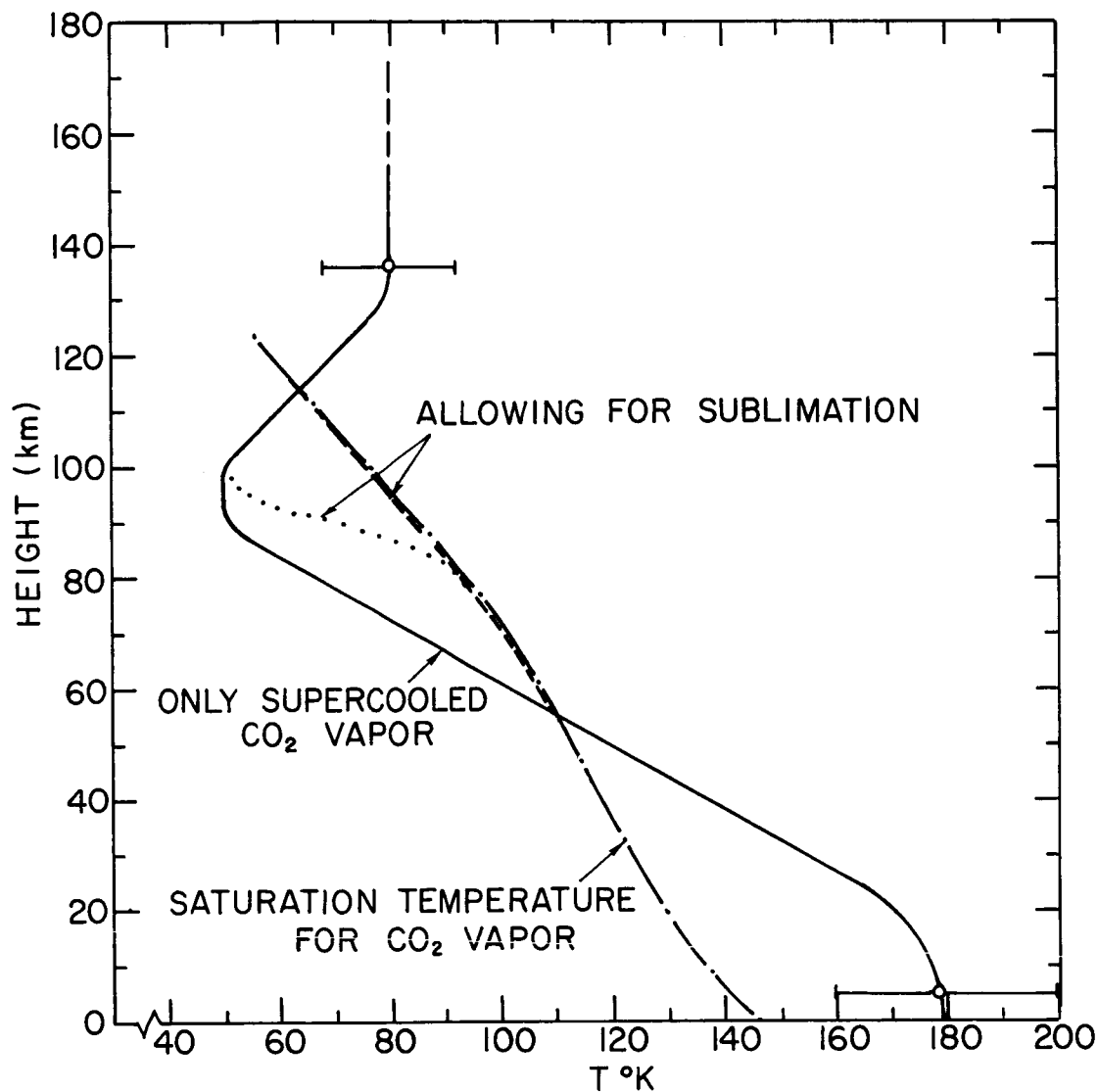


FIG. 3. TEMPERATURE VS ALTITUDE ABOVE ELECTRIFIED LAYER AT THE TIME OF IMMERSION INTO OCCULTATION. The limiting ion loss process in the main ionospheric layer is assumed to be $O^+ + CO_2 \rightarrow CO + O_2^+$ with a rate coefficient of $10^{-9} \text{ cm}^3/\text{sec}$.

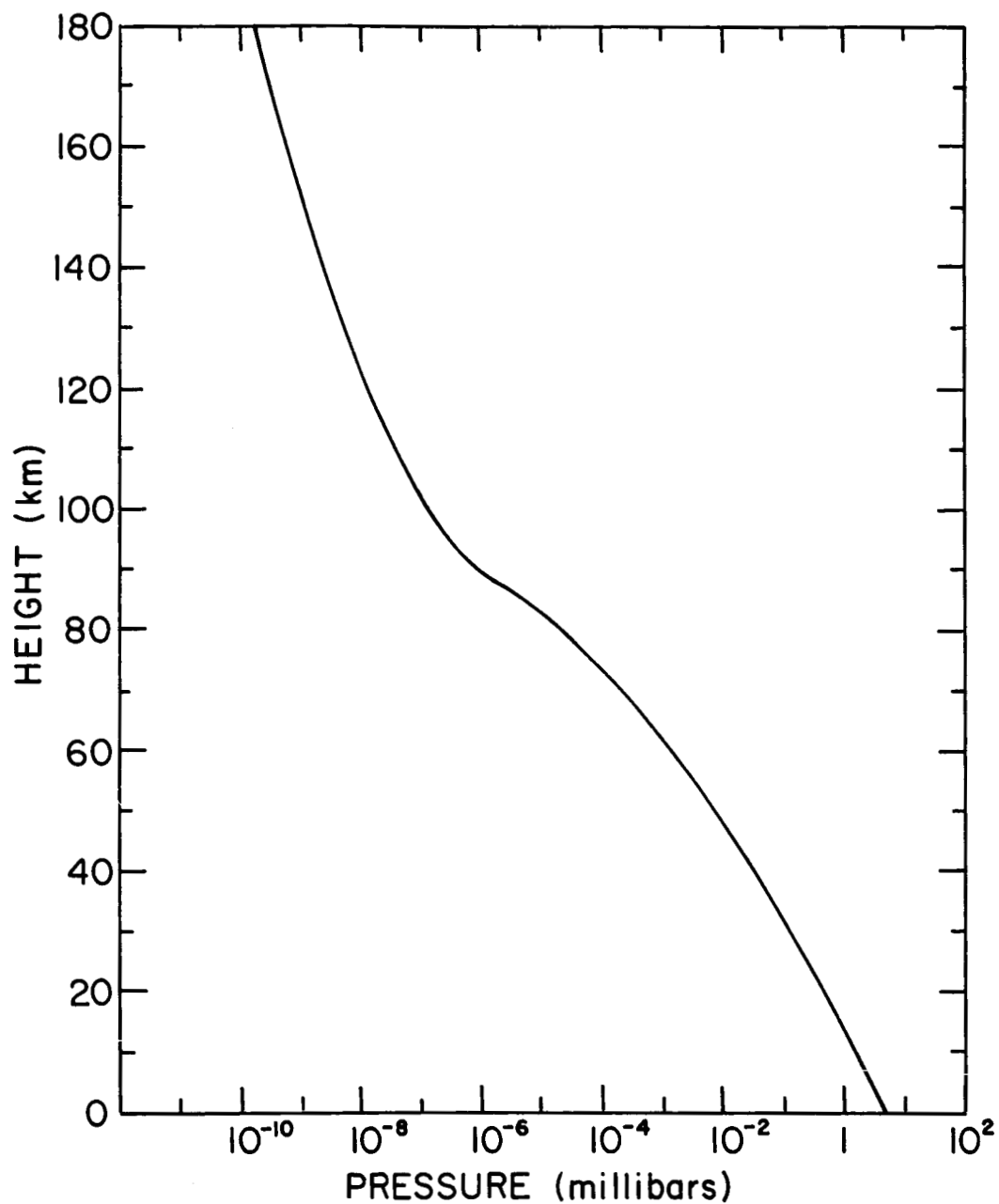


FIG. 4. PRESSURE VS ALTITUDE ABOVE ELECTRIS AT THE TIME OF IMMERSION INTO OCCULTATION. The limiting ion loss process in the main ionospheric layer is assumed to be $O^+ + CO_2 \rightarrow CO + O_2^+$ with a rate coefficient of $10^{-9} \text{ cm}^3/\text{sec}$.

Decreasing (or increasing) k_1 by one order of magnitude has the effect of increasing (or decreasing) $n(\text{CO}_2)$ in the upper atmosphere of Mars by a factor of 10. However, this would not change our identification of (5) as the most likely "bottleneck" reaction, particularly if the other rate coefficients have a similar temperature dependence.

For the model shown in Fig. 2, the dissociation region lies between 70 and 80 km altitude. The carbon dioxide dissociation is largest at unity optical depth. Thus, the height of this region depends on the $n(\text{CO}_2)$ profile and reducing k_1 tends to increase the dissociation height and vice versa. It should also be pointed out that a complete solution for the number density distribution in the dissociation region would have to account for O_2 , Eqs. (3) and (4), although this would have little effect at higher altitudes.

The number density profiles shown in Fig. 2 were derived assuming that CO_2 is in diffusive equilibrium. This approach is justified in Fig. 5, where it is shown that for this model the diffusion time constant at all heights is smaller than the time constants for loss of CO_2 .

The full drawn curve in Fig. 3 shows a model for the temperature profile above Electris at the time of occultation immersion. Two criteria were used in order to construct a reasonable temperature profile below 30 km altitude:

1. The number density profile should have the observed scale height of about 9 km in this region.
2. Very low temperatures are needed between 30 and 100 km altitude in order to provide the low densities at ionospheric heights.

In the ionosphere, the temperature was found from the electron scale height under the assumption that the electrons and ions are in temperature equilibrium with the neutral constituents.

The atmospheric region between 30 and 100 km altitude is not well known since there is neither enough neutral molecules nor electrons at these heights to affect the radio-phase path measurably. Heat flows into this region by conduction from the thermosphere (about 100 to 140 km) and the lower atmosphere. In addition, there is heating due to absorption of solar ultraviolet and X-rays. At the temperature minimum

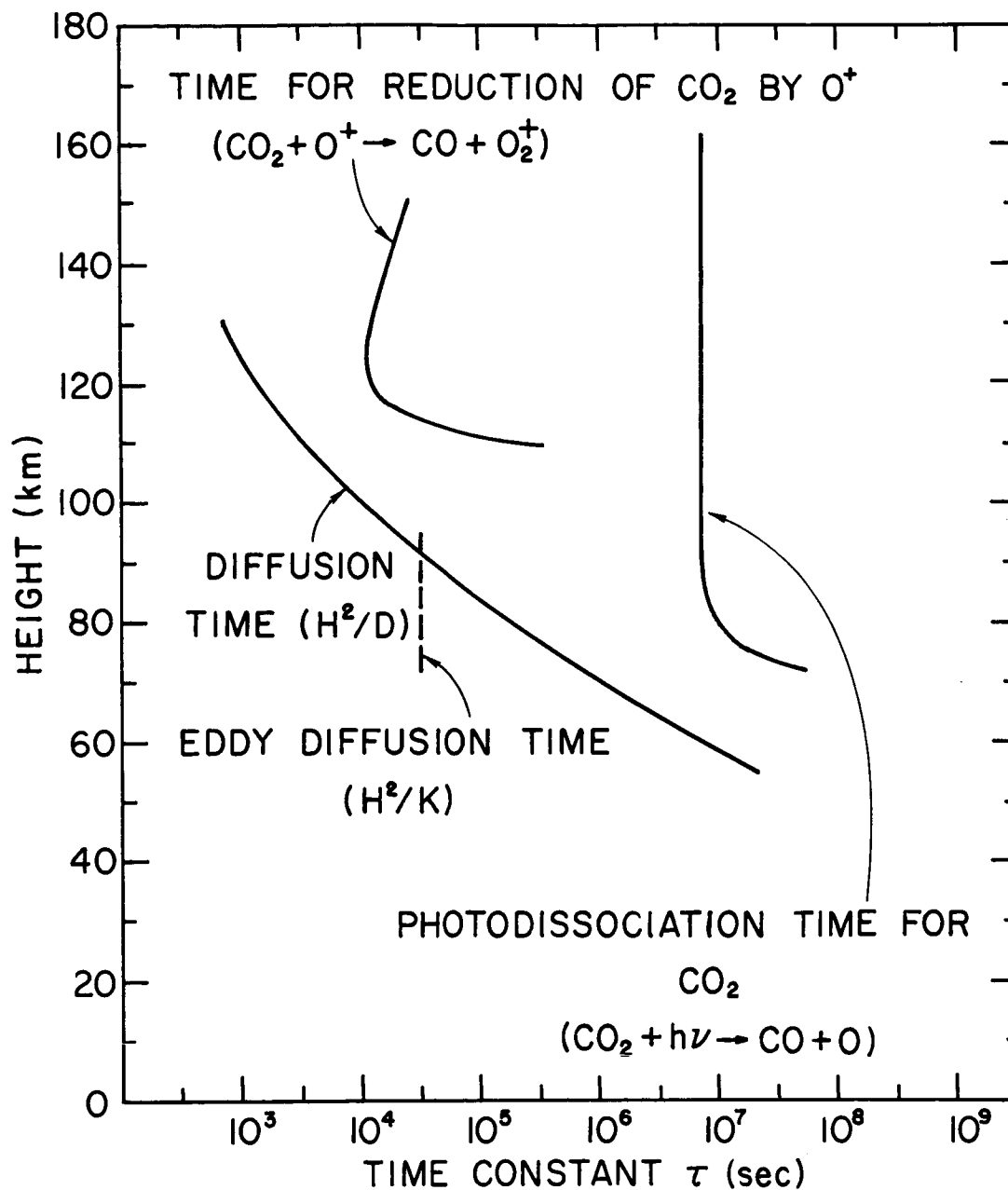


FIG. 5. TIME CONSTANTS VS ALTITUDE FOR DIFFERENT PROCESSES INVOLVING CO_2 . The figure applies to the carbon dioxide - atomic oxygen model discussed in the text.

(or minima) which is expected to lie between 50 and 100 km altitude, the heat can only be removed by radiation. Carbon monoxide, atomic oxygen, and carbon dioxide are all important infrared radiators in this region.

The actual temperature profile between 30 and 100 km altitude can at this time only be obtained by inference. We know, for instance, that the integral of T^{-1} from the surface to the top side of the ionosphere is uniquely determined by the surface and ionospheric densities and temperatures. With the value for k_1 adopted here, the CO_2 number density must decrease by 11 orders of magnitude from the surface to 120 km altitude, which indicates that the scale height and temperature at intermediate levels of the atmosphere are substantially lower than the values obtained near the surface. This result is further substantiated by the ionospheric electron density profile, where the sharp gradients on the bottom side suggest a temperature even lower than the 80 °K found for regions above the peak.

The minimum temperature at which all CO_2 would remain a gas, under stationary conditions, is also indicated in Fig. 3. Carbon dioxide may be supercooled in a broad region around 80 km altitude. If there are suitable sublimation nuclei present, one would expect formation of dry ice particles. This mechanism has been suggested as a source of particle layers in the Martian atmosphere [Ref. 14].

Let us consider briefly the effect on the number density and the temperature profiles if CO_2 does sublimate. This can be done in a simplified manner by changing the temperature and the number density in such a way that the pressure profile for the CO_2 gas remains the same. By increasing the temperature until the gas is just saturated and removing the excess CO_2 gas molecules by permitting them to produce particles, then one obtains the stippled temperature and number density profiles shown in Figs. 3 and 2 respectively. An intermediate situation is also indicated by the dotted temperature curve in Fig. 3. For this case, the CO_2 gas is allowed to remain supercooled around 100 km altitude where collisions occur less frequently, sublimation nuclei may be rare, and the diffusion time is shorter. To what extent the CO_2 gas actually sublimates depends on the characteristic time constant for this process.

Because of lack of pertinent data, this time constant has not been included in Fig. 5.

We have also considered the case that carbon dioxide neither is supercooled nor forms particles. For this case, we find that the reaction rate coefficient k_1 would have to be about two orders of magnitude smaller than given above in order to be consistent with the peak electron number density obtained from the occultation measurements.

Although more work remains to be done, it appears that the low temperature obtained from the occultation data are consistent with the conclusion that CO_2 predominates at ground level. This is because CO , O , and CO_2 are more effective in radiatively cooling the atmosphere, particularly near the dissociation region, [Refs. 15, 18, and 19] than is N_2 , which was previously believed to be the main constituent. It should also be noted that the solar irradiance at Electris was nearly one order of magnitude lower than the solar constant at the earth at this time of sunspot minimum, which explains the low temperature gradients in the thermosphere above Electris.

The electron number density profile shown in Fig. 1 indicates a minor layer at approximately 100 km. The neutral density in this region is too low for it to be like the terrestrial E layer. However, this region coincides quite closely with the level where the maximum O^+ production by solar ultraviolet (200-800 Å) occurs, so that it should be identified as an F1 layer. While O^+ production is maximum here, O_2^+ [formed by reaction (5)] may actually be the principal ion since its lifetime is greater than O^+ below about 115 km.

V. OTHER MODELS FOR THE UPPER ATMOSPHERE

Reaction (6) is another possibility for the O^+ "bottleneck" reaction at the peak of the main layer. For a rate coefficient of about 10^{-11} cm^3/sec , one finds that there would have to be of the order of 10^8 oxygen molecules per cm^3 at 120 km in order to account for the ion loss rate. Ten percent O_2 appears to be too high an abundance to be acceptable, based on our preliminary studies. Also the spectroscopic observations indicate that there is less than 0.1 percent O_2 in the lower atmosphere of Mars [Ref. 8]. However, the O_2 distribution in the Martian atmosphere is complicated by the fact that atomic oxygen, produced by the photodissociation of carbon dioxide, associates to form O_2 . Thus molecular oxygen might still be of some importance for the loss of O^+ , at least in the lower portion of the ionosphere.

If reaction (7) is assumed to be the principal loss process in the main layer, one finds that there would have to be of the order of 10^9 N_2 molecules/ cm^3 at 120 km altitude. Using this figure, together with the number density and temperature profiles shown in Figs. 2 and 3, we find an N_2 profile which remains higher than the CO_2 profile all the way down to where mixing dominates over diffusion. By analogy with the earth's atmosphere this transition is expected to take place at a total number density level of about 3×10^{12} molecules/ cm^3 . However, since CO_2 is the principal constituent in the lower atmosphere of Mars, we conclude that reaction (5) is much more likely than reaction (7) for the loss of O^+ .

Another approach one might take in designing atmospheric models for Mars is to assume that the atomic oxygen formed through photodissociation of CO_2 and O_2 (and also through dissociative recombination of CO_2^+ , CO^+ and O_2^+ with electrons) is being lost. If this for instance could take place to such an extent that CO^+ or CO_2^+ became the principal ion, then the main ionospheric layer would be of a Chapman type. The temperature on the top side, corresponding to a molecular mass of 28 or 44, would be 140 and 220 °K, respectively.

However, we have not found any plausible mechanism which would remove atomic oxygen to the extent that $n(O)$ would remain lower than $n(CO_2)$ or $n(CO)$ all the way up to the exosphere. Thermal escape of O would, for instance, require exospheric temperatures in excess of 1000 °K [Ref. 14]. We have also considered the possibility that solar protons might sweep away atomic oxygen in the exosphere of Mars, but find that collisions do not occur frequently enough to carry away what can be supplied through upwards diffusion. It therefore appears that we are left with atomic oxygen as the most probable principal constituent in the upper atmosphere of Mars.

VI. ATMOSPHERIC MASS DENSITY

Figure 6 illustrates how the atmospheric mass density varies with altitude for any of the oxygen models (Bradbury ionosphere) where reaction (5), (6), or (7) is the controlling loss mechanism. The mass density at 120 km altitude would be of the order of 3×10^{-14} g/cm³ for these models. However, the mass density could be a factor of 30 higher at this altitude if the main layer were a Chapman layer. This difference in interpretation results in the uncertainty indicated at 120 km altitude in Fig. 6. Even so, the density must still be lower than would follow from extrapolation utilizing the new surface values, and very substantially lower than was previously contemplated.

Since gravity is about 62 percent lower on Mars, it was thought that the atmospheric mass density would decrease much slower with altitude than on the earth. Thus, nearly independent of the atmospheric pressure assumed near the surface, it was proposed that above 30 to 50 km altitude the density would become larger than the density at the same altitudes in the earth's atmosphere. However, the occultation data analysis reveals quite a different state of affairs. The low temperatures in the upper and lower atmosphere, together with the larger molecular mass, compensate for the lower gravity in controlling the scale height. Hence, the mass density profile does not cross over the corresponding profile for the earth's atmosphere but remains several orders of magnitude lower all the way up to the exosphere which begins at about 140 km (Bradbury layer) or 180 km (Chapman layer), where the mean-free-path equals the scale height.

The question of how the mass density varies with altitude in the Martian atmosphere has also some practical implications. In order to place a satellite in orbit around Mars, it is important to know how the atmospheric density varies with height. Most previous models required an orbit altitude of up to several thousand kilometers for a satellite with a lifetime of 50 years, but the occultation measurements show that a few hundred kilometers should be sufficient.

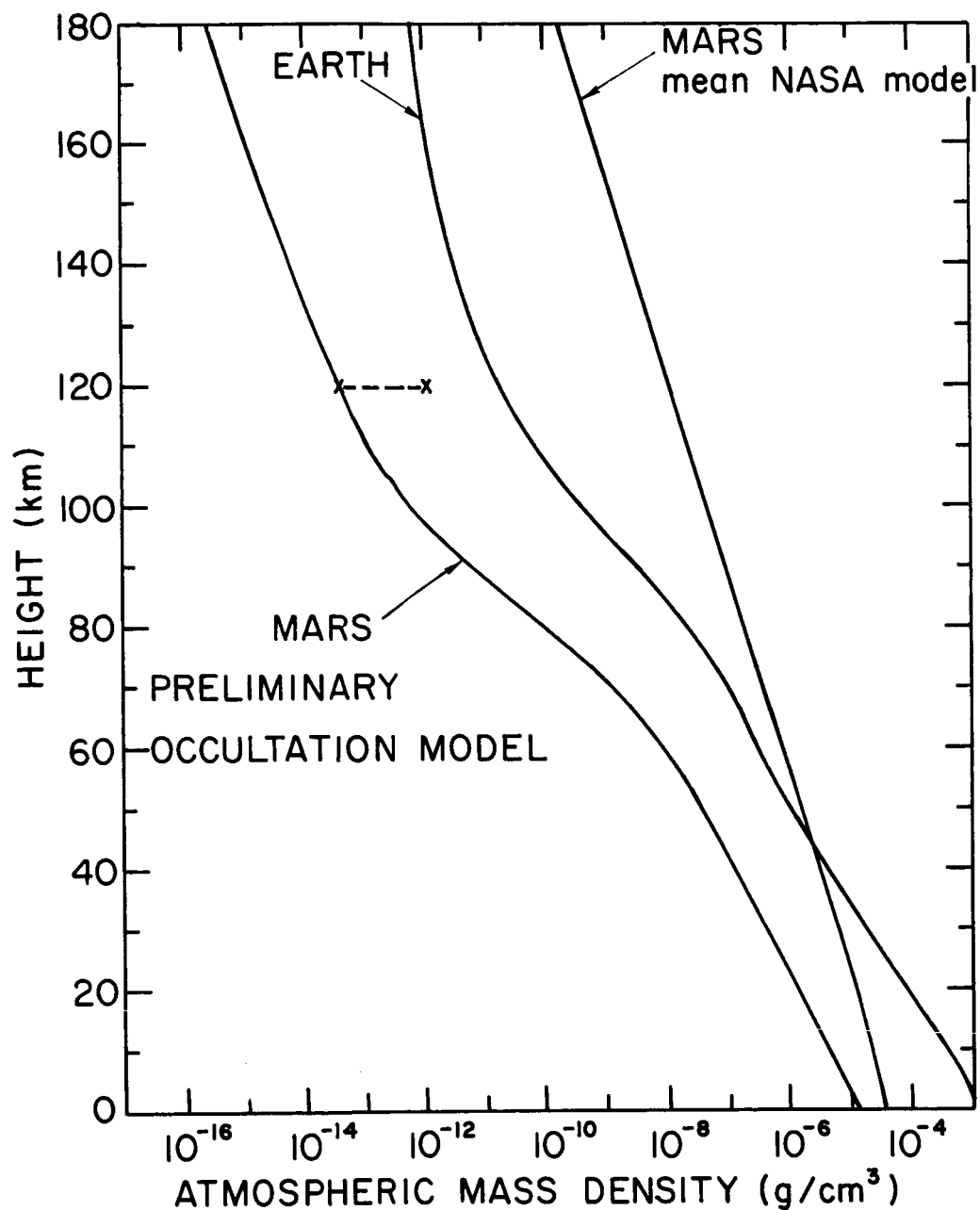


FIG. 6. ATMOSPHERIC MASS DENSITY VS ALTITUDE ABOVE ELECTRIS AT THE TIME OF IMMERSION INTO OCCULTATION. The NASA engineering model shown here was used for planning purposes prior to the Mariner IV mission.

VII. CONCLUDING REMARKS

The principal difficulty in attempting to identify the critically important ion loss process in the main ionospheric layer of Mars is the uncertainty in reaction rate coefficients, particularly with respect to their temperature dependence. Based on present information, we believe that the atmospheric model presented in the figures [the atomic oxygen - carbon dioxide model of the upper atmosphere, based on Eq. (5)] is more likely than the other possibilities that were considered, although revisions may be necessary when more information on temperature dependence becomes available. This favored model requires only CO_2 near the surface, and thus we neither need any minor constituents to understand the atmospheric processes, nor can we use the occultation results to identify minor constituents. Other considerations indicate that N_2 and Ar may be present [Refs. 6-9] but as discussed previously, CO_2 must be the principal atmospheric constituent.

F. S. Johnson [Ref. 5] has also derived an atmospheric model based on the ionospheric occultation results of Mariner IV. From a comparison, however, it appears that there are several important differences in treatment and conclusions from those of Fjeldbo et al [Ref. 4] and the work reported here. Johnson does not discuss which reactions are responsible for the loss of O^+ in the main layer. If the measurement by Fehsenfeld et al [Ref. 17] of the rate coefficient for reaction (5) applies, Johnson's atmospheric model is not compatible with the measured ionospheric profile. Numerically, his model would yield an O^+ density nearly three orders of magnitude lower than the electron density measured at 120 km, and his F2 peak would be at 140 km instead of at 120 km.

Johnson's conclusion that there is no thermosphere on Mars also differs from our work, which suggests a positive temperature gradient below the peak of the ionosphere to help transport heat downward to regions where it may be radiated more efficiently. Also, we do not agree that his lapse rate of $5^\circ\text{K}/\text{km}$ in the troposphere over Electris is compatible with the experimental results or with what would be expected from radiative exchange theory for a carbon dioxide atmosphere at this latitude and

season. However, his work agrees with ours in the fundamental points concerning the preponderance of atomic oxygen in the upper atmosphere and the conclusion that the main ionospheric layer is of the Bradbury (F2) type. Johnson has also made important contributions with regard to the geologic history of the Martian atmosphere, a subject which is not treated here.

REFERENCES

1. A. J. Kliore, D. L. Cain, G. S. Levy, V. R. Eshleman, G. Fjeldbo, and F. D. Drake, "Occultation Experiment: Results of the First Direct Measurement of Mars's Atmosphere and Ionosphere," Science, 149, 1965, pp. 1243-1248.
2. A. J. Kliore, D. L. Cain, F. D. Drake, V. R. Eshleman, G. Fjeldbo, and G. S. Levy, "Preliminary Results of the Mariner IV Occultation Measurements of the Atmosphere of Mars," Proc. Caltech - JPL Lunar and Planetary Conference, Pasadena, California, September 13-18, 1965.
3. D. L. Cain, F. D. Drake, V. R. Eshleman, G. Fjeldbo, A. J. Kliore, and G. S. Levy, "A Summary of Preliminary Results of the Mariner IV Radio Occultation Experiment," Proc. Ionospheric Research Committee of the Avionics Panel, Advisory Group for Aerospace Research and Development, NATO, Rome, Italy, September 21-25, 1965.
4. G. Fjeldbo, V. R. Eshleman, A. J. Kliore, D. L. Cain, G. S. Levy, and F. D. Drake, "Preliminary Results of the Mariner IV Radio Occultation Measurements of the Upper Atmosphere of Mars," Proc. Caltech - JPL Lunar and Planetary Conference, Pasadena, California, September 13-18, 1965.
5. F. S. Johnson, "Atmosphere of Mars," Science, 150, 1965, pp. 1445-1448.
6. R. M. Goody, "The Atmosphere of Mars," Weather, 12, 1957, pp. 3-15.
7. L. D. Kaplan, G. Munch, and H. Spinrad, "An Analysis of the Spectrum of Mars," Astrophys. J., 139, 1964, pp. 1-15.
8. T. C. Owen and G. P. Kuiper, "A Determination of the Composition and Surface Pressure of the Martian Atmosphere," Comm. Lunar Planetary Lab., No. 32, 1964.
9. D. E. Evans, "Ultraviolet Reflectivity of Mars," Science, 149, 1965, pp. 969-972.
10. P. J. Nawrocki and R. Papa, Atmospheric Processes, Prentice Hall, Inc., New Jersey, 1963.
11. Y. Mintz, Appendix 8 from The Atmospheres of Mars and Venus, prepared by W. W. Kellogg and C. Sagan, Pub. 944, National Academy of Sciences - National Research Council, Washington, D.C., 1961.
12. G. Fjeldbo and V. R. Eshleman, "The Bistatic Radar-Occultation Method for the Study of Planetary Atmospheres," J. Geophys. Res., 70, 1965, pp. 3217-3225.

13. G. Fjeldbo, V. R. Eshleman, O. K. Garriott, and F. L. Smith, III, "The Two-Frequency, Bistatic, Radar-Occultation Method for the Study of Planetary Ionospheres," J. Geophys. Res., 70, 1965, pp. 3701-3710.
14. J. W. Chamberlain, "The Upper Atmospheres of the Planets," Astrophys. J., 136, 1962, pp. 582-593.
15. R. B. Norton, "A Theoretical Study of the Martian and Cytherian Ionospheres," NASA Tech. Note TN d-2333, National Aeronautics and Space Administration, Washington, D.C., July 1964.
16. H. Rishbeth and D. W. Barron, "Equilibrium Electron Distributions in the Ionospheric F2-Layer," J. Atmos. Terr. Phys., 18, 1960, pp. 234-252.
17. F. C. Fehsenfeld, A. L. Schmetlekopf, and E. E. Ferguson, J. Chem. Phys., to be published, private communication.
18. D. R. Bates, "The Temperature of the Upper Atmosphere," Proc. Phys. Soc., B64, 1951, pp. 805-821.
19. M. B. McElroy, J. L'Ecuyer, and J. W. Chamberlain, "Structure of the Martian Upper Atmosphere," Astrophys. J., 141, 1965, pp. 1523-1535.